

Estimating Recharge Rates with Analytic Element Models and Parameter Estimation

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Abstract

Quantifying the spatial and temporal distribution of recharge is usually a prerequisite for effective ground water flow modeling. In this study, an analytic element (AE) code (GFLOW) was used with a nonlinear parameter estimation code (UCODE) to quantify the spatial and temporal distribution of recharge using measured base flows as calibration targets. The ease and flexibility of AE model construction and evaluation make this approach well suited for recharge estimation. An AE flow model of an undeveloped watershed in northern Wisconsin was optimized to match median annual base flows at four stream gages for 1996 to 2000 to demonstrate the approach. Initial optimizations that assumed a constant distributed recharge rate provided good matches (within 5%) to most of the annual base flow estimates, but discrepancies of >12% at certain gages suggested that a single value of recharge for the entire watershed is inappropriate. Subsequent optimizations that allowed for spatially distributed recharge zones based on the distribution of vegetation types improved the fit and confirmed that vegetation can influence spatial recharge variability in this watershed. Temporally, the annual recharge values varied >2.5-fold between 1996 and 2000 during which there was an observed 1.7-fold difference in annual precipitation, underscoring the influence of nonclimatic factors on interannual recharge variability for regional flow modeling. The final recharge values compared favorably with more labor-intensive field measurements of recharge and results from studies, supporting the utility of using linked AE-parameter estimation codes for recharge estimation.

Introduction

Recharge, defined here as water that crosses the water table, depends on a wide variety of factors (e.g., vegetation, precipitation, climate, topography, geology, and soil type), making it one of the most complex and uncertain hydrologic parameters to quantify. Understanding the spatial and temporal distribution of recharge is often essential for effective ground water flow modeling.

While a few investigators (e.g., Edmunds et al. 2002; Jyrkama et al. 2002) have considered both the spatial and temporal variability of recharge at a watershed scale, ground water modelers often ignore recharge variability and assume a single, constant recharge value for a watershed. Although a single constant value may be adequate for long-term simulation of regional ground water flow systems (Juckem et al. in press), it may be inappropriate for predictions where small-scale or detailed time-dependent flowpath delineation is required (Jyrkama et al. 2002). In particular, spatial and temporal recharge variability has important implications for site-scale ground water budget calculations, flowpath calculation, nutrient cycling, and contaminant transport.

Although others (e.g., Hunt et al. 2000; Kelson et al. 2002) have linked parameter estimation and analytic element (AE) codes, and it is well known that flow models can be calibrated to estimate regional recharge rates (e.g., Martin and Frind 1998; Varni and Usunoff 1999), to our knowledge, this work represents the first use of an AE

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code with a parameter estimation code to quantify temporally and spatially distributed recharge rates.

Methods

The approach described here assumes that the stream is the only outlet for flow and there is no loss of water from the ground water system via evapotranspiration; thus, the amount of water entering the ground water system via recharge must equal the amount that exits the system via base flow in the stream. Consequently, if the flow system is steady state, base flow is estimated from field measurements, and if the contributing area to the stream is known, recharge can be estimated from measurements of base flow. Previous researchers (e.g., Rutledge and Daniel 1994; Mau and Winter 1997) used stream hydrographs to estimate basinwide recharge rates. These analyses provide a single average annual recharge rate for an entire basin. The methods require that the ground water contributing area for the stream gage be known, but accurate delineation of the gage's ground water contributing area can be difficult. Hydrogeologists routinely use the topographically delineated surface watershed as a proxy for the ground watershed, even though surface watershed and ground watershed boundaries are usually not coincident (e.g., Sander 1971; Hunt et al. 1998; Winter et al. 2003), particularly in areas of low or hummocky topography. Furthermore, hydrograph separation techniques do not provide spatially distributed recharge distributions at the subbasin scale. Ground water models provide a means to estimate

the temporal and spatial variability in ground water recharge values from streamflow data without a priori knowledge of the ground water-contributing area and can be used to explore and help quantify regional controls (e.g., vegetation) on recharge distribution.

In this study, a two-dimensional AE code for ground water flow (GFLOW, Haitjema and Kelson 1994; Haitjema 1995) was used with a nonlinear parameter estimation code (UCODE, Poeter and Hill 1998) to illustrate how linked AE-parameter estimation codes can be used to estimate recharge using median base flows from stream gages as calibration targets. The two codes were used to estimate recharge during 1996 to 2000 in the Trout Lake basin, a relatively small (120 km²) watershed located in north-central Wisconsin (Figure 1). The AE model was optimized to match estimated median annual base flows at four stream gages within the watershed (Figure 1). Accurate estimation of annual recharge from base flow measurements is predicated on the assumptions of no loss of ground water to deeper aquifers and an annual steady-state flow system.

Loss of ground water to a deep aquifer (thus not captured by a single-layer AE model) is not expected in the study area. The Trout Lake basin comprises 30 to 50 m of unconsolidated sands atop essentially impermeable Precambrian metamorphic and igneous bedrock (Okwueze 1983; Attig 1985). The hydraulic conductivity of the sands is significantly larger than that of the bedrock such that loss of ground water from the sands into the underlying bedrock is negligible.

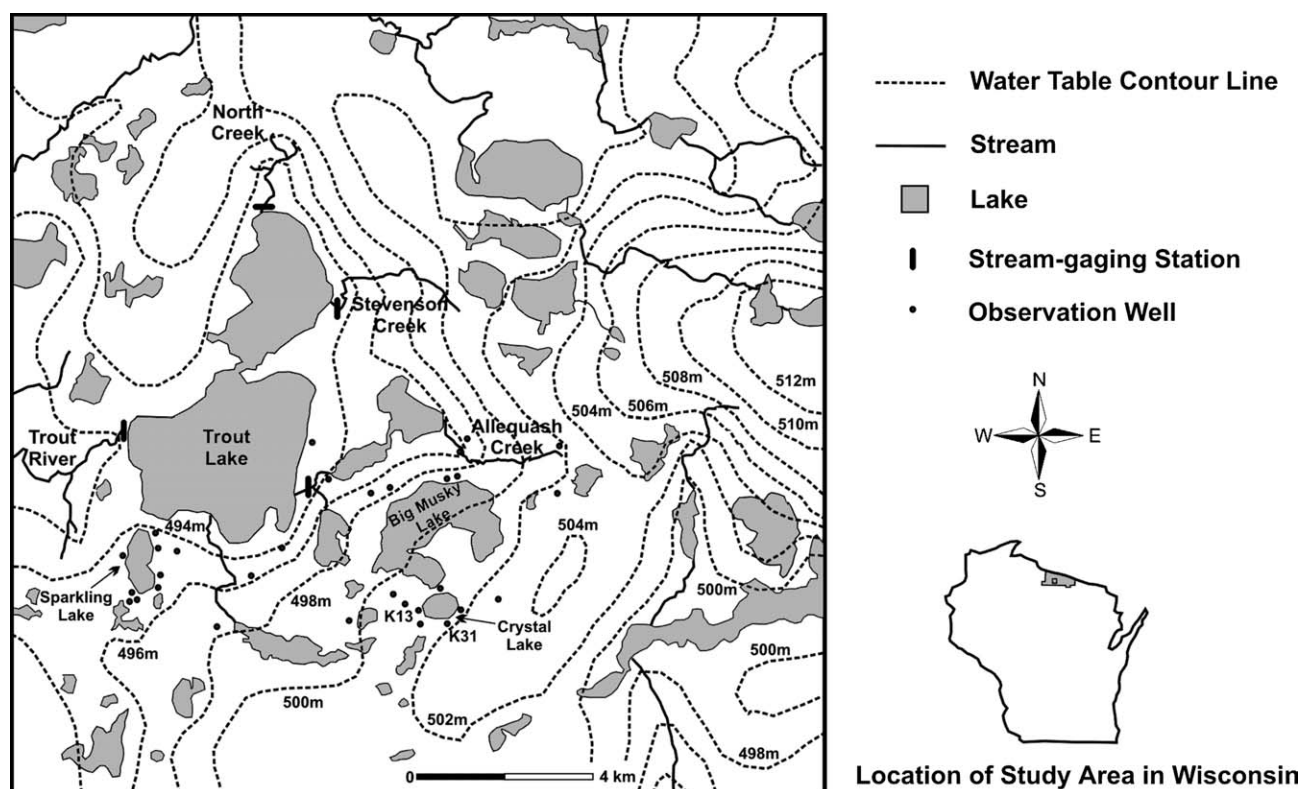


Figure 1. Map of the Trout Lake study area in northern Wisconsin, showing elevation of the water table above mean sea level and locations of stream-gaging stations and observation wells. Water levels in well K13 and K31 are shown in Figure 2.

Biweekly lake-level and bimonthly well water measurements at select lakes and wells within the basin (Figure 2) suggest that a steady-state assumption is reasonable on an annual basis for 1996 to 2000, with the exception of 1998, when lake and well water levels decreased (Figure 2). The implications of this decrease on recharge estimation are discussed in a subsequent section.

Although there is a network of observation wells in the watershed (Figure 1), head targets were not used in this recharge estimation exercise because (1) many of the wells were near head-specified surface water bodies limiting the usefulness of the target (Hunt 2002); (2) the datasets from many of the wells did not have extensive coverage for the period of this study; and (3) the objective of this work was on the basinwide flux distribution, which is more easily evaluated using flux targets, which cover larger portions of the basin than point measurements of head. Heads were indirectly included, however, as the model used in this work is an extension of previous models constructed for the basin (e.g., Hunt et al. 1998; Pint et al. 2003; Hunt et al. 2003a,b) where head targets were formally included in the calibration. Initially, the model was optimized to a constant recharge value for the basin; subsequent optimizations incorporated spatially distributed recharge zones based on the basin's vegetation types in order to evaluate the improvement in fit that resulted from the additional parameterization.

AE models were first introduced by Strack and Haitjema (1981a, 1981b) and further developed and discussed by Strack (1989, 1999), Haitjema (1995), and Mitchell-Bruker and Haitjema (1996), among others. AE models are increasingly used for regional ground water flow modeling studies (e.g., Hunt et al. 1998; Bakker et al. 1999; Hunt et al. 2000; Kelson et al. 2002). The existing, two-dimensional AE (GFLOW) model of Hunt et al. (1998) for the Trout Lake basin was modified for the purposes of this study. GFLOW is well suited for this application due to the ease and flexibility of AE model construction, evaluation, and refinement, as well as the

ability to conjunctively simulate stream-aquifer interaction, and the independence of scale inherent in AE methods.

Parameter estimation models were introduced in the 1970s (e.g., Cooley 1977) and have recently been applied more routinely in ground water modeling studies (e.g., Doherty 1994; Sun et al. 1995; Boonstra and Bhutta 1996; McLaughlin and Townley 1996; Poeter and Hill 1997; Kitanidis 1997; Weiss and Smith 1998). The linking of parameter estimation approaches to AE models was first implemented using only head targets (Power and Barnes 1993) and then using head and flux data (e.g., Hunt et al. 2000; Kelson et al. 2002).

UCODE (Poeter and Hill 1998) performs parameter estimation using nonlinear regression. The nonlinear regression problem is solved by minimizing a weighted least-squares objective function with respect to the parameter values using a modified Gauss-Newton method. UCODE is a "universal" code that can be linked to any application model that can be run in batch mode and uses text files for input and output. UCODE optimizes the results of the application model, determining parameter values that provide a quantified best fit between the simulated values and the observed values that form the calibration targets.

Study Site and Model Design

The Trout Lake basin is sparsely populated with glaciated terrain and rolling upland hills covered with a mixed temperate forest of deciduous and coniferous trees interspersed among kettle lakes (Figure 1). The kettles are remnants of the last continental glaciation ~10,000 years ago. Glaciers scoured the impermeable Precambrian metamorphic and igneous bedrock surface and deposited 30 to 50 m (Okwueze 1983; Attig 1985) of unconsolidated sand and coarse till atop the bedrock as they receded northward. Trout Lake is the major lake in the basin and is drained by the Trout River to the west and fed by four streams (Figure 1). Streamflows were recorded continuously from 1996 to 2000 at gages at the outlet of Trout Lake and on Allequash, Stevenson, and North Creeks (Figure 1).

For this example, the original GFLOW model of Hunt et al. (1998) was modified by adding additional surface water features, refining the discretization and location of a subset of the stream and lake elements (Hunt et al. 2003b), and use of additional recharge zones (Figures 3A and 3B). An AE model has no explicit perimeter boundary, but element properties can be varied depending on its location in the near field (in the area of interest) or far field. Those elements within and near the boundaries of the Trout Lake watershed are modeled as near field (dashed line in Figures 3A and 3B) while those around the periphery are treated as far field (dotted line in Figures 3A and 3B). The near-field features are represented with significantly more detail, with specific attention to the location and geometry of the elements. The far-field features are represented more coarsely and include only major surface water features, which essentially control

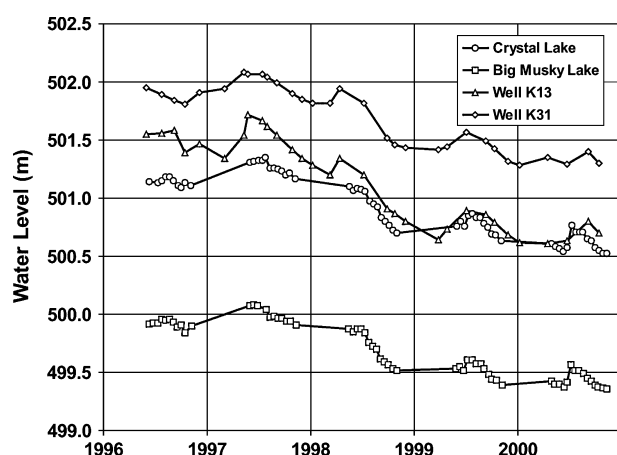


Figure 2. Measured water levels in select lakes and wells within the Trout Lake watershed for 1996 to 2000. Water elevations at the beginning and end of each year are roughly comparable, except for 1998 during which water elevations appreciably declined.

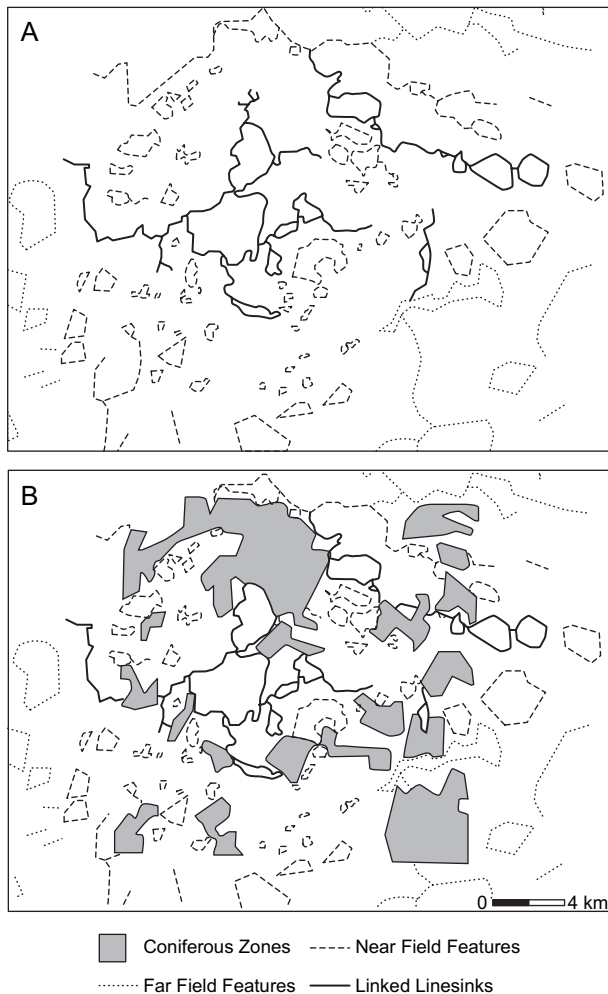


Figure 3. (A) Design of the AE model for the Trout Lake watershed (B) with the location of the coniferous recharge zones (shaded in gray). Unshaded areas represent deciduous recharge zones. The dashed line elements are near-field features. The dotted line elements are far-field features. The solid line elements are linked linesinks that are modeled in GFLOW using the conjunctive surface water–ground water solution.

flow toward or away from the near field and help define the hydrologic “boundaries” of the near-field system.

Streams are represented by strings of linesinks that simulate ground water extraction or stream water infiltration. Lakes are modeled by linesink strings along their perimeters, where most of the lake-aquifer interaction takes place. The linesinks (streams and lakes) were not assumed to be in perfect communication with the aquifer. In the model, resistance was assigned based on the type of the surface water feature, as recommended by Haitjema (1995). Water velocity through the stream bottoms is faster than through the lake bottoms; the lake bottom sediments are thus slightly finer grained than the stream bottom sediments. Based on field observations, stream bottoms were assumed to have a 0.3-m-thick layer of fine sands with a vertical hydraulic conductivity of 3 m/d, roughly one-third of the aquifer’s hydraulic conductivity; lake bottoms were assumed to have a 0.3-m-thick layer of finer silty sand with a vertical hydraulic conductivity of 0.3 m/d. Consequently, lakes were assigned a resistance of 1 d and streams were assigned a resistance of 0.1 d.

The near-field stream and lake linesinks that comprise the Trout River system were linked into an interconnected stream network (solid lines in Figures 3A and 3B), and the conjunctive surface water–ground water solution in GFLOW was used to provide base flow estimates at the location of each of the four gages in the basin. Stream elevations and lake stages were initially assigned based on the 1992 USGS topographic map of the area but were adjusted in the model on an annual basis based on lake-level measurements at select lakes within the basin from 1996 to 2000. The aquifer base was set to 450 m above sea level, resulting in an average aquifer thickness of ~50 m (Okwueze 1983; Attig 1985). A hydraulic conductivity of 8.64 m/d was assumed based on fieldwork (Dripps 2003) and previous modeling studies in the basin (Hunt et al. 1998; Champion 1998; Pint et al. 2003; Hunt et al. 2003a and b). The GFLOW model was optimized to estimate the annual terrestrial recharge rate for 1996 to 2000 by using the median annual value of the measured daily stream base flows as calibration targets. Annual lake recharge was specified for all lakes in the watershed for each year as the difference between the annual measured precipitation and the estimated annual lake evaporation. The annual precipitation was measured at a precipitation gage at the Noble Lee Airport, ~11 km southwest of Trout Lake. The Noble Lee Airport Station is the closest station with a complete precipitation record for the model period of interest (1996 to 2000). A relatively evenly spaced network of six tipping bucket rain gages was installed in the Trout Lake watershed during the summer of 1999 to provide data on rainfall intensity, duration, distribution, and timing within the basin. The data from these gages were compared to that collected at the Noble Lee Airport Station. Based on the gage comparison, precipitation was relatively spatially homogeneous across this area, with annual differences among the gages <5%. As such, the data from the Noble Lee Airport Station were considered representative of precipitation across the basin for 1996 to 2000. The annual lake evaporation was estimated using an energy budget technique and daily meteorological data collected on Sparkling Lake (Kratz 2002) (location shown in Figure 1). The difference between the regional recharge rate and the lake recharge rate was accounted for in GFLOW by adjusting areal recharge inhomogeneities within each lake. The resulting recharge rates for the lakes were (in cm/year) 45 for 1996, 18 for 1997, –19 for 1998, 24 for 1999, and 21 for 2000.

The nonlinear regression technique used a least-square formulation for an objective function, defined in terms of the residual between simulated results and field observations, and assigned weights. A weight is assigned to each calibration target to designate the relative importance of a particular observation. Median annual base flow estimates (Table 1) served as the only targets for the recharge optimizations and were calculated using the PART method (Rutledge 1993, 1998). Weights were assigned to each base flow target based on uncertainties for the daily discharge measurements for each gage (coefficient of variation in Table 1). The goodness of fit between the observed and simulated base flows is reflected by the sum of the squared weighted residuals (SOSWR). The

Table 1
Observed (PART) vs. Simulated Base Flows (in m³/d) for the Different
GFLOW-UCODE Recharge Optimizations

	Observed Base Flow		Simulated Base Flows			
	PART Base Flow	Coefficient of Variation	Single Value ¹	% Error from Observed	Zoned by Land Cover ²	% Error from Observed
1996						
Trout	123,283	0.057	125,028	1.4	122,926	−0.3
Allequash	31,870	0.064	32,779	2.9	31,913	0.1
Stevenson	8363	0.085	8353	−0.1	8416	0.6
North	10,795	0.075	10,330	−4.3	10,775	−0.2
SOSWR				0.59		0.01
1997						
Trout	111,682	0.052	112,062	0.3	111,464	−0.2
Allequash	33,325	0.067	31,479	−5.5	31,121	−6.6
Stevenson	8241	0.085	8131	−1.3	8190	−0.6
North	9,514	0.075	9891	4.0	10,113	6.3
SOSWR				0.99		1.69
1998						
Trout	73,359	0.05	73,606	0.3	73,628	0.4
Allequash	23,698	0.063	23,839	0.6	23,592	−0.4
Stevenson	6725	0.087	6550	−2.6	6688	−0.6
North	7954	0.075	7633	−4.0	7963	0.1
SOSWR				0.39		0.01
1999						
Trout	71,117	0.05	74,759	5.1	72,288	1.6
Allequash	21,271	0.058	21,900	3.0	20,638	−3.0
Stevenson	5443	0.085	5515	1.3	5673	4.2
North	6798	0.075	5929	−12.8	6603	−2.9
SOSWR				4.24		0.76
2000						
Trout	73,961	0.052	78,324	5.9	77,070	4.2
Allequash	22,791	0.052	22,991	0.9	21,980	−3.6
Stevenson	6462	0.092	5983	−7.4	6261	−3.1
North	7600 ³	0.092	6712	−11.7	7546	−0.7
SOSWR				3.50		1.26

¹The Single Value numbers are the simulated base flows assuming a constant value of recharge.

²The Zoned by Land Cover values are the simulated base flows for the land cover optimization.

³Estimated from a partial record by comparison to historical records and streamflows measured at the other gages in the basin.

lower the SOSWR value, the better the match to the calibration targets.

Results

Spatial Distribution of Recharge Rates

Initial optimizations assumed a constant value of recharge for each simulated year and one recharge zone for the entire model domain. Overall, there is reasonable agreement between the observed and simulated base flows for the 5-year period (Table 1), although discrepancies at individual gages in some years are as large as 12.8% (Table 1). With the exception of 1997, the fits for Stevenson Creek, Allequash Creek, and the Trout River are consistently better (−7.4% to 5.9% difference from the observed base flows) than those for North Creek (−12.8% to 4% difference from the observed base flow) (Table 1). Excluding 1997, the model consistently underestimates base flow at North Creek (Table 1). The disparities

at North Creek suggest that either (1) the model inadequately represents or omits some of the hydrologic features within this subbasin and/or (2) the system properties are not as homogeneous as represented in the model.

With regard to the first possible explanation for the disparities, all surface water features shown on the USGS topographic map, except wetlands, are included in the model. Lake evaporation is explicitly incorporated into the model, but the evapotranspiration of ground water from wetlands is not. Riparian vegetation can transpire appreciable quantities of near-surface ground water, resulting in a lower base flow. Indeed, it has been noted that near-stream ground water levels can be below stream stage during the growing season in this area (Hunt et al. in press). Although wetland evapotranspiration may affect the base flow targets, it cannot explain why North Creek's base flow is commonly underestimated by the model.

With respect to the second possible explanation, well logs, slug test data, and field mapping (Dripps 2003) show that the Trout Lake watershed is covered by

relatively homogeneous glacial outwash sands and gravels. There is no physical evidence to suggest that the hydraulic conductivity values for the near-surface glacial sediments that cover the North Creek subwatershed are markedly different from those of the rest of the basin. This leaves spatial differences in recharge rates as the most plausible explanation.

Additional optimizations were run that incorporated spatially distributed recharge zones based on the basin's vegetation types in an attempt to improve the model fit and help quantify regional controls (e.g., vegetation) on the spatial recharge distribution. All optimizations were performed using the same annual base flow targets. Thus, the total recharge for both sets of optimizations was similar (0% to 3.3% difference in net recharge applied to the basin between the sets of optimizations among the 5 years). Results from a soil water balance (SWB) model (Dripps 2003) and from IBIS-2, a terrestrial biosphere model (Foley et al. 1996; Kucharik et al. 2000), which were used to calculate recharge in the study area over the same 5-year period, suggest that the recharge distribution is primarily controlled by differences in vegetation (Dripps 2003). The AE model linked to a parameter estimation code provided a means to assess and help quantify the role of vegetation on the spatial recharge variability in the watershed.

Land cover affects the distribution of recharge as each cover type can be expected to have a different interception capacity, evapotranspiration rate, and shading effect, all of which influence recharge. The basin is almost completely forested, consisting of a patchwork of coniferous and deciduous stands. Coniferous stands are believed to have higher annual recharge rates compared to deciduous stands owing to (1) lower evapotranspiration rates due to the conifer's leaf structure and (2) the shading effects of the conifer canopy during the winter months that reduce ablation and inhibit snow loss to evaporation (Dripps 2003). The digital WISCLAND land cover distribution (Gurda 1994) was used as a basis to divide the land area into two recharge zones, coniferous forests and the remaining land area (predominantly deciduous forests). The coniferous recharge areas were specified in the AE model using recharge inhomogeneities (Figure 3B), and the model was reoptimized using two recharge zones.

With the exception of 1997, the addition of the land cover-based recharge zones appreciably improved the

model fit, as reflected by the decrease in the SOSWR (Table 1). In particular, the discrepancies for the North Creek gage were reduced from a range of -12.8% to 4% to a range of -3.1% to 6.3% (Table 1). Although it might be expected that additional parameterization should enhance the fit (given the addition of degrees of freedom), it remains unclear why the zonation did not improve the fit in 1997. One possible explanation is that the measurements may have been more uncertain in 1997, as indicated by the anomalous year-to-year change (i.e., the Allequash target increased compared to 1996, but all other targets decreased). Indeed, the conifer recharge zone was applied to areas away from North Creek (Figure 3); thus, there may have been calibration trade-offs between fitting North Creek and the other targets.

Although small discrepancies still remain between the simulated and observed base flows, the results suggest that spatial variability in recharge is present and is likely influenced by the vegetation distributions. The recharge rate in coniferous areas was found to be 3 to almost 10 cm (7% to 55%) greater than the recharge rate in deciduous areas, depending on the year (AE Deciduous Zone and AE Coniferous Zone columns in Table 2). The estimated recharge rates compare favorably with the range of annual recharge estimates calculated by a land surface/soil/vegetation model (IBIS-2), an SWB model, and field-based, stream hydrograph analyses (Dripps 2003) (Table 2). The recharge values for the IBIS-2 and SWB models (Table 2) represent the range of annual recharge values calculated by these grid-based models for all cells within the Trout Lake basin. The range results from differences in land cover and soil texture among the model cells. In both models, land cover was identified as the major driver for the calculated spatial recharge variability, with soil type serving as a secondary contributor (Dripps 2003). The higher recharge values within the range are indicative of the coniferous stands; the lower values are indicative of the deciduous stands. The recharge values in the Field column in Table 2 represent the range of annual recharge values calculated by analyzing daily streamflow records at the four stream-gaging stations within the Trout Lake basin (Figure 1) using a recession curve displacement method (Rutledge 1993, 1998).

Spatial variability in recharge caused by the coniferous and deciduous stands reflects the interplay between

Table 2
Comparison of GFLOW-UCODE Recharge Estimates (listed as AE Deciduous Zone and AE Coniferous Zone) with Other Models and Field Data from Dripps (2003)

Year	Precipitation (cm)	Annual Recharge Estimates (cm)				
		AE Deciduous Zone	AE Coniferous Zone	IBIS-2	SWB	Field
1996	98.0	44.0	50.9	36.1–43.9	33.2–43.6	23.4–50.6
1997	77.8	44.1	47.0	35.6–42.8	31.2–40.8	23.8–44.1
1998	57.2	31.2	34.4	18.6–20.4	12.0–16.2	18.1–37.6
1999	80.5	17.5	27.1	15.7–25.6	11.0–29.1	14.7–31.3
2000	79.0	22.3	32.0	21.8–30.0	17.4–28.9	13.1–27.7

the climatic drivers (e.g., precipitation) and the terrestrial drivers (e.g., land cover) on the recharge process. Recharge not only is dependent on the amount of precipitation but also is a function of the timing of precipitation events. Rainfall intensity can affect recharge by affecting overland flow. Also, the antecedent moisture at the time of a precipitation event is influenced by the type of vegetation present. The observed differences in recharge rates, in this instance, can be primarily attributed to differences in soil moisture deficits, which differ at the time of particular recharge events between the two vegetation types. These differences in soil moisture deficit develop during the growing season, when different plant communities have different transpiration rates and different rates of removing moisture from the soil.

Temporal Recharge Variability

The GFLOW-UCODE results show that recharge varies interannually in the Trout Lake watershed (Table 2). With the exception of 1998, the annual GFLOW-UCODE recharge results are comparable to the field measurements and annual recharge estimates from other models of the basin for 1996 to 2000 (Table 2). The decline in water elevations in lakes and wells during 1998 (Figure 2) indicates that the steady-state flow assumption used in this research is not valid for 1998 and likely accounts for the discrepancy in recharge estimates between the AE-optimization approach and the other models. That is, the decline in water elevations indicates a decrease in the volume of ground water held in storage within the basin. This stored water supplemented measured 1998 base flows and reflected water recharged in previous years rather than water recharged in 1998. Base flow only represents ground water in the stream, and as such does not discriminate between recharge from 1998 and from recharge stored from previous years and released with the decline in storage. This technique and, for that matter, techniques that use base flow as a surrogate for recharge require a steady-state system for accurate annual recharge estimation. It is difficult to ascertain annual changes in storage from streamflow records alone, and thus it is essential that head measurements (from wells and/or surface water bodies) are available and used to test the validity of the requisite steady-state assumption (Figure 2). By assuming that the system is at steady state and all base flow was derived from recharge that occurred during that year, the AE optimization for 1998 consequently yields simulated annual recharge estimates that are too large. For example, the 1998 modeled recharge values of 31.2 and 34.4 cm/year for deciduous and coniferous zones, respectively, are larger than the values computed by IBIS-2 and SWB (Table 2).

While disparities in recharge rates for 1998 stand out, differences in other years between the recharge estimates from the AE optimization and other models may be partly caused by transient effects as well (Figure 2). In hydrogeologic settings in which aquifers and streams respond more quickly to temporal variations in recharge (i.e., smaller storage coefficients, higher transmissivities, shorter distances between surface water), transient effects will be less influential and the successive steady-state

Analytic Element Model optimization technique will give better results (Haitjema 1995). The steady-state assumption is a definite limitation to using this technique, as was the case in 1998.

The interannual variability highlights the importance of transience in hydrologic systems and has important implications for ground water flow modeling and ecological analyses. Between 1996 and 2000, the highest estimated annual recharge rate was >2.5 times the lowest estimated annual recharge rate. During this same period, the highest annual precipitation was only 1.7 times the lowest annual precipitation (Table 2). This underscores that there is not necessarily a direct correlation between precipitation and recharge, and that temporal variability in recharge can be significant in regional ground water budget calculations even for undeveloped watersheds. Although annual recharge is roughly correlated with annual precipitation, years with comparable precipitation, like 1997 and 2000, can have markedly different values of recharge (the estimated recharge for 1997 is almost double that for 2000, Table 2). The variability in annual recharge highlights the importance of temporal controls like antecedent moisture, timing of precipitation, and snowmelt on the recharge process. Although the exact mechanisms are not well understood, it is thought that recharge was significantly higher in 1997 due to (1) a significantly larger snowpack and snowmelt event; (2) smaller soil moisture deficits prior to major storm events; and (3) more precipitation during the spring and fall seasons when the vegetation was devoid of foliage, transpiration was at a minimum, and the soil moisture levels were consequently higher and more conducive to recharging the ground water system (Dripps 2003). Regardless of uncertainty in the drivers, the interannual variation is appreciable and is likely expected in other midwestern basins that have similar climatic regimes and geologic settings. When variations over short time periods are of concern, water resource planners and ground water flow modelers should include a range of recharge that reflects annual variability rather than long-term averages for planning and modeling purposes. A series of steady-state flow models used in this study provide a quick and practical means to quantify the spatial distribution of recharge; this type of insight, in turn, should lead to better water resource management.

Conclusions

A two-dimensional AE code for ground water flow (GFLOW) was used with a parameter estimation code (UCODE) to illustrate how linked AE-parameter estimation codes can be used to estimate the spatial and temporal distribution of recharge using steady-state models and measured annual base flows as calibration targets. The ease and flexibility of AE model construction, evaluation, and refinement coupled with automated optimization make this approach well suited for recharge estimation. To demonstrate the approach, a linked GFLOW-UCODE model was used to estimate annual recharge rates for 1996 to 2000 in the Trout Lake watershed by matching median annual base flows measured at four stream gages.

The final recharge values compared favorably with other methods that included field measurements of recharge results from a terrestrial biosphere model (IBIS-2) and an SWB model. This general agreement illustrates the viability of using linked AE-parameter estimation codes for recharge estimation and highlights the magnitude and significance of spatial and temporal recharge variability.

Including differences in vegetation community appreciably improved the model calibration, suggesting that vegetation can be an important driver for the spatial recharge variability. Interannual variability in recharge was greater than spatial variability and was likely the result of a combination of climatic (e.g., timing and amounts of precipitation) and terrestrially based drivers (e.g., land cover, soil moisture content). The spatial and interannual variability of recharge highlights the potential for heterogeneity and year-to-year transience in hydrologic systems, and the importance of considering transience in analyzing ground water systems.

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